

occurred heterogeneously, as previously discussed [3], the redox state gradually changed and sulfidation weakened accordingly. The gravitational segregation of metal-sulfide and silicate was enhanced at the later stage of accretion and eventually caused the fractionation of REE, which was reflected in EL5 and EL6.

**References:** [1] Shinotsuka K. et al. (1995) *Meteoritics*, 30, 694–699. [2] Kallemeyn G. W. and Wasson J. T. (1986) *GCA*, 50, 2153–2164. [3] Kong P. and Ebihara M. (1996) *Meteoritics & Planet. Sci.*, 31, A72. [4] Crozaz G. and Lundberg L. L. (1995) *GCA*, 59, 3817–3831.

**STABLE ISOTOPES IN ALLAN HILLS 84001: AN ION MICRO-PROBE STUDY.** J. M. Eiler<sup>1</sup>, J. W. Valley<sup>2</sup>, and E. M. Stolper<sup>1</sup>, <sup>1</sup>Geological and Planetary Sciences, California Institute of Technology, Pasadena CA 91125, USA (eiler@expet.gps.caltech.edu), <sup>2</sup>Geology and Geophysics, University of Wisconsin, Madison WI 53713, USA.

The temperature of formation of carbonates in ALH 84001 provide a test of the hypothesis that they contain features produced by ancient martian life [1], and contributes to our understanding of the global budgets of martian volatiles. We have determined the micrometer-scale distribution of O and C isotope ratios in carbonate concretions, enstatite, and SiO<sub>2</sub> in ALH 84001 by ion microprobe as a means of constraining the range of possible temperatures of fluid-rock interaction [2]. The principal results are: (1) Carbonate δ<sup>18</sup>O is 9.5–20.6‰, SMOW, significantly expanding the range from bulk analysis. The magnesite rim is higher in δ<sup>18</sup>O than Ca-rich (0.07 ≤ X<sub>Ca</sub> ≤ 0.13) cores. (2) Variations in δ<sup>18</sup>O occur over length scales as small as 50 μm. (3) Interiors of fractured orthopyroxene that hosts carbonate are homogeneous to within ±1‰. (4) Secondary SiO<sub>2</sub> has a δ<sup>18</sup>O of 20.4‰. These data are inconsistent with mutual O isotope equilibrium among carbonate and silicate minerals and indicate either fine-scale mineral-mineral equilibrium at a range of temperatures ≤300°C or a failure to attain mineral-mineral equilibrium. Mineralogical evidence for disequilibrium supports the latter interpretation. Our δ<sup>18</sup>O values for carbonates have been confirmed by two studies [3], and extended to compositions of carbonate not found in our sample.

Ranges in δ<sup>18</sup>O among carbonates in ALH 84001 and other SNC meteorites have in the past been interpreted as temperature constraints based on fluid-carbonate equilibrium (likely in any case) and models for the isotopic composition of martian volatiles [e.g., 4]. Constraints using our data and such models are: (1) Variations in temperature during precipitation from a salt-free, juvenile, H<sub>2</sub>O-CO<sub>2</sub> fluid; the temperature range must be 0°–175°C. (2) Variations in the temperature of precipitation from an aqueous fluid of constant salinity and (unknown) δ<sup>18</sup>O; the temperatures must have ranged up to at least 50°C. If salinity and temperature covaried, these limits are relaxed and the maximum temperature could have been somewhat lower. (3) Changes in chemical composition of the fluid (e.g., CO<sub>2</sub>/H<sub>2</sub>O ratio) at constant temperature; temperature less than 300°C. (4) Precipitation from two or more fluids of different δ<sup>18</sup>O; no temperature limits. No one of these models can be preferred using existing data. Consideration of recent data by Gilmore et al. and Leshin et al. in terms of these models result in relatively minor changes from these limits. Other models appear generally plausible but are contraindicated by the existing data (e.g., variation in δ<sup>18</sup>O due to Rayleigh distillation from CO<sub>2</sub>-rich fluid, suggested by [3], violates the covariation of δ<sup>18</sup>O and δ<sup>13</sup>C determined for ALH 84001 carbonates by partial dissolution experiments [5] and violates the expected relationship between the bulk isotopic composition of carbonate and the range determined by ion probe).

We conclude that no rigorous temperature constraint can be made using existing data, and that the best estimate derives from comparison with terrestrial analogies for the processes that led to the observed metastable mineralogy, isotopic variability, and oscillatory chemical zoning. The best terrestrial analogies of which we are aware are (1) diagenetic siderites, (2) carbonate weathering products of ultramafic rocks, and (3) carbonates precipitated from aqueous fluids in subgreenschist facies crustal rocks. The temperature limits indicated by these analogies are uncertain, but imply a maximum of ~300°C and typically occur only at near-surface temperatures. This interpretation is consistent with (but does not indicate) biological activity during carbonate precipitation. Of more general significance, this in-

terpretation indicates that carbonates in ALH 84001 are an example of CO<sub>2</sub> dissolved in aqueous fluids being sequestered in carbonate minerals in the martian crust. While the extent of this process on Mars is unknown, this geochemical cycle is among the most important interactions between the crust and atmosphere on Earth.

**References:** [1] McKay et al. (1996). [2] Valley et al. (1997) *Science*. [3] Gilmore et al. and Leshin et al. (1997) *LPS XXVIII*, verbal communication. [4] Clayton and Mayeda (1988). [5] Romanek et al. (1994).

**THE SAGA OF MASKELYNITE IN SHERGOTTY.** A. El Goresy<sup>1</sup>, B. Wopenka<sup>2</sup>, M. Chen<sup>3</sup>, and G. Kurat<sup>4</sup>, <sup>1</sup>Max-Planck-Institut für Kernphysik, Postfach 103980, D-69029 Heidelberg, Germany (goresy@selket.mpi-hd.mpg.de), <sup>2</sup>Department of Earth and Planetary Sciences, Washington University, St. Louis MO 63130, USA, <sup>3</sup>Institute of Geochemistry, Academia Sinica, Guangzhou, China, <sup>4</sup>Naturhistorisches Museum, P.O. Box 417, A-1014 Vienna, Austria.

Maskelynite was first discovered in the Shergotty meteorite by Tschermack [1]. Until recently, it has been considered to be a diaplectic glass that was formed by shock-induced solid-state transformation of plagioclase during impact events on the parent planet [2–5]. Maskelynite in Shergotty was reported to have retained its stoichiometric plagioclase composition and preshock chemical igneous zoning [2–5]. However, El Goresy et al. [6] recently presented evidence that the material thought to be maskelynite in Peace River and Tenham (both L6) formed by local melting of plagioclase, followed by Na and K metasomatism. This finding prompted our careful investigation of the alleged maskelynite in Shergotty.

The nature of the material described as maskelynite in Shergotty is assessed by the presence and/or absence of the following petrologic features: (1) preshock cleavage, (2) preshock cracks formed during cooling after crystallization of the igneous rocks, (3) shock-induced fractures that penetrate both pyroxene and maskelynite, (4) original chemical zoning, and (5) evidence of melting of feldspar and pyroxene and melt injections in fractures. Petrologic (optical microscopy, SEM, EMPA) and laser Raman microprobe (LRM) spectroscopic investigations were conducted on individual selected areas of grains.

We recognize two types of pseudo-maskelynites:

**“Maskelynite” Grains of Type 1:** The overwhelming majority of the grains (more than 97%) are smooth with no cleavage, no contraction cracks, and no shock-induced fractures. These grains display offshoots of smooth “maskelynitic” material into fractures of the adjacent pyroxene, thus attesting that a liquid was injected in the fractures after passage of the shock wave. The grains also contain schlieren of molten pyroxene. The lack of cleavage, cracks, and specifically shock-induced fractures in the grains and their long offshoots indicate that these textural features were erased in a melting event. Those smooth type 1 grains are always surrounded by numerous radiating cracks that have extensively shattered the neighboring pyroxene, titanomagnetite, and troilite. These cracks are indicative of volume increase of the glass that was quenched from the melt under high pressure and expanded upon pressure release [6]. The Raman spectra confirm that the grains are not crystalline: The spectra of the “maskelynite” grains cannot be distinguished from those obtained on the nonstoichiometric glass in Tenham, and they are characteristic of amorphous silicate glass. Systematic EMPA profiles across smooth type 1 grains reveal distinct chemical signatures in that they do not possess the typical zoning of igneous feldspars. The curves for An and Ab contents are flat, do not display a systematic complementary behavior, and the compositional range is extremely narrow (An<sub>57.6</sub>Ab<sub>42.3</sub>Or<sub>0.7</sub> to An<sub>55</sub>Ab<sub>45</sub>Or<sub>0.9</sub>), indicating homogenization. In many cases the chemical composition of individual grains follows the same trend: decrease or increase in both An and Ab contents with constant Or contents. In a few grains the profiles are indicative of the presence of chemically coherent residual fragments (An<sub>56.4</sub>Ab<sub>46</sub>Or<sub>0.7</sub> to An<sub>49.7</sub>Ab<sub>48.1</sub>Or<sub>1.2</sub>) of unknown nature.

**“Maskelynite” Grains of Type 2:** Those are longitudinal grains with sharp boundaries that display reduced birefringence, remnant albite twinning, rare cleavage, and fractures. Compositional EMPA profiles display well-behaved zoning with An-rich cores (An<sub>53.6</sub>Ab<sub>46.6</sub>Or<sub>0.9</sub> to An<sub>50.1</sub>Ab<sub>47.2</sub>Or<sub>1.2</sub>) and Ab-rich rims (An<sub>45.2</sub>Ab<sub>50.8</sub>Or<sub>1.9</sub> to An<sub>40.4</sub>Ab<sub>47.1</sub>Or<sub>2.3</sub>).